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# The evolution of the southern margin of the East European Craton based on seismic and potential field data

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**Abstract**

This paper presents an integrated geophysical study of the southern margin of the East European Craton (EEC) in the Karpinksy Swell-North Caucasus area. It presents new interpretations of deep refraction and wide-angle reflection "deep seismic sounding" (DSS) data as well as conventional seismic and CDP profiling and new analyses of potential field data, including three-dimensional gravity and magnetic modelling. An integrated model of the physical properties and structure of the Earth's crust and, partially, upper mantle displays distinct features that are related to tectonic history of the study area. The Voronezh Massif (VM), the Ukrainian Shield and Rostov Dome (RD) of the EEC as well as the Donbas Foldbelt (DF), Karpinsky Swell (KS), Scythian Plate (SP) and Precaspian Basin (PCB) constitute the geodynamic ensemble that developed on the southern margin of the continent Baltica. There proposed evolutionary model comprises a stage of rifting during the middle to late Devonian, post-rift extension and subsidence during Carboniferous–early Permian times (synchronous with and related to the southward displacement of the Rostov Dome and extension in a palaeo-Scythian back-arc basin), and subsequent Mesozoic and younger evolution. A pre-Ordovician, possibly Riphean (?), mafic magmatic complex is inferred on a near vertical reflection seismic cross-section through the western portion of the Astrakhan Dome in the southwest part of the Precaspian Basin. This complex combined with evidence of a subducting slab in the upper mantle imply the presence of pre-Ordovician (Riphean?) island arc, with synchronous extension in a Precaspian back-arc basin is suggested. A middle Palaeozoic back-arc basin ensemble in what is now the western Karpinsky Swell was more than 100 km to the south from its present location. The Stavropol High migrated northwards, dislocating and moving fragments of this back-arc basin sometime thereafter. Linear positive magnetic anomalies reflect the position of associated faults, which define the location of the eastern segment of the Karpinsky Swell. These faults, which dip northward, are recognised on crustal DSS profiles crossing the Donbas Foldbelt and Scythian Plate. They are interpreted in terms of compressional tectonics younger than the Hercynian stage of evolution (i.e., post-Palaeozoic).


*Keywords:* Scythian Plate; Karpinsky Swell; Crustal structure; East European Craton

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## 1. Introduction

The tectonic history, and the processes that controlled them, of the main tectonic units along the southern margin of the East European Craton (EEC), is a key objective for understanding Peri-Tethyan geology. The Black Sea Basin, Azov Sea Basin, Scythian Plate (SP) and Precaspian Basin (PCB), as well as the Dnieper–Donets Basin (DDB) and its inverted southeasternmost portion, which is known as the Donbas Foldbelt (DF), and the deformed sedimentary successions comprising what is now the Karpinsky Swell (KS) are the main tectonic units that developed, were reworked, and/or accreted to the EEC since the middle to late Palaeozoic (Fig. 1).

A fixist tectonic view of the evolution of the southern margin of EEC traditionally dominated until the end of 1970s in the former Soviet Union (Letavin, 1980). New conceptual models of Mesozoic to Cainozoic tectonic assembly for the area under discussion, developed in a plate tectonic framework, have been proposed during the last decade (e.g. Nikishin et al., 1996, 1998a,b). Stovba et al. (1996) proposed a model of intra-cratonic rifting for the DDB in the late Devonian and post-rift subsidence during Carboniferous to Cretaceous times based on reprocessing and on reinterpretation existing seismic data in the Ukraine. They also demonstrated the rift origin of the DF (Stovba and Stephenson, 1999). A new review of the deep structure and evolution of PCB was done by Brunet et al. (1999). Zonenschain et al. (1990) suggested that a Scythian orogenic belt lies just to the south from the southern boundary of the EEC and is the result of collision of several small microcontinents.

Nevertheless, the tectonic setting and mechanisms of origin of the SP in pre-Mesozoic time are poorly known. The geometry and tectonic position of KS are also poorly understood. Traditionally, the latter is considered to be the northern portion of the SP (Letavin, 1980). Conversely, Volozh et al. (1999) regarded the KS as an element of a super-long intracratonic rift, comprising from the northwest to

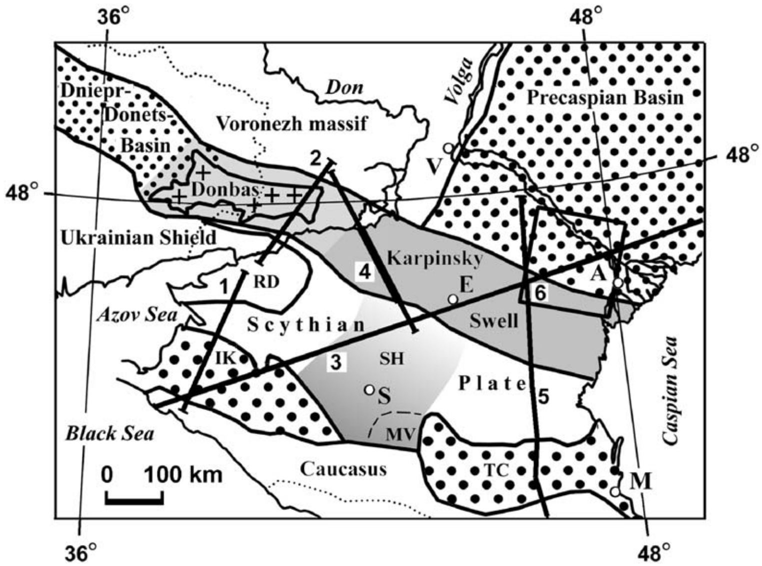


Fig. 1. Tectonic setting and location map of regional seismic profiles discussed in the text. Dotted line indicates the boundary between Ukraine and Russia (west) and between Russia and Georgia (south). Black straight lines are the seismic profiles: 1 = Ilskaya–Leningradskaya, 2 = Bataisk–Milyutinskaya, 3 = Krasnodar–Emba, 4 = Morozovsk–Manich (thick portion of line indicate location of CMP cross-section presented on Fig. 4a,b), 5 = Volgograd–Nakhichevan. Square frame, numbered 6, is the area of deep CDP lines within the Astrakhan dome shown in Fig. 6a. Crosses show the eroded part of the Donbas Foldbelt. Cities: V = Volgograd, E = Elista, S = Stavropol, A = Astrakhan, M = Makhachkala. Selected tectonic units: RD = Rostov Dome, StH = Stavropol High, MV = Mineralnie Vody Dome, IKB = Indol–Kuban foredeep, TCB = Tersk–Caspian foredeep.

southeast the DDB, DF, KS and the Tuarkyr zone. The last of these is just to the southeast of the Caspian Sea running to the Kopet-Dag orogen.

The present study presents a re-evaluation of the tectonic setting and history of the southern margin of the EEC mainly from seismic, gravity and magnetic data. New seismic data from the study area—namely the international DOBRE seismic project in southern Ukraine (Stephenson, 1997; DOBREfraction '99 Working Group, 2003) and the combined refraction/wide-angle reflection and near-vertical and converted wave deep seismic experiment along the Morozovsk–Manich–Elbrus regional profile in southern Russia (Kostyuchenko et al., 2001) are also considered, providing new constraints on 2-D and 3-D gravity and magnetic modelling. The gravity analysis paper of Yegorova et al. (2004) can be considered as a companion paper.

## 2. Geophysical data and their interpretation

### 2.1. Seismic studies

Seismic studies of the crust began in the study area in the mid-1950s. Specifically, several deep seismic sounding (DSS) profiles across the eastern portion of Donbas and neighbouring parts of the EEC and KS were carried out in 1971–1973 by "Dniepergeofizika" and "Spetsgeofizika" and, in 1974–1975, by the "Central Geophysical Trust" of the former USSR. Conventional refraction studies and shallow and middle-deep (5–10 s TWT) CDP profiles dominate in the eastern KS and adjacent regions of the PCB. Three deep CDP seismic reflection lines (20 s TWT) were acquired in that area (Brodsky et al., 1993, 2000; Brodsky and Voronin, 1994). Since 1990, GEON acquired DSS data based on the registration of refracted, wide-angle reflected and converted waves. Combined DSS-CDP observations along the Morozovsk–Manich–Elbrus regional profile were begun by GEON in 1999, with the aim of studying the structure and architecture of the crust of the southern slope of the Voronezh Massif (VM), the central portion of the KS, the Stavropol High (StH) of the SP and the pre-Caucasus foreland. Fig. 1 shows the location of the main seismic profiles that are discussed in the text.

Figs. 2–7 present velocity models of the internal structure of the crust and, in part, the upper mantle for the study area along profiles located in Fig. 1. The results of different seismic studies are in different stages of completion. Old seismic data have not been reprocessed and remodelled but new geological interpretations of the previously determined velocity structures are presented.

Fig. 2a shows the Ilskaya–Leningradskaya DSS profile across the western portion of the SP near the Azov Sea (profile 1 in Fig. 1; Volvovsky and Volvovsky, 1975). Fig. 2b shows the Bataisk–Mylyutinskaya DSS profile across the DF (2 in Fig. 1; Sollogub et al., 1978). Each of these incorporates refraction data aimed at elucidating the velocity structure of the upper crust as well as wide-angle reflection data for the middle and lower crust. Fig. 2c is a composite interpretation of the velocity structure. The refraction boundary with P-wave velocity 5.6 km/s and greater marks the top of basement, which is considered to be heterogeneous in age. Borehole data from the Rostov Dome (RD) indicate that velocities in the range 5.9–6.2 km/s are characteristic of basement that is of Archean to early Proterozoic age and that velocities of 5.6–5.7 km/s are typical of the Palaeozoic successions. There are no geological observations that can be directly related to the observed velocities of 6.4 km/s and more at the top of the basement of the Indol–Kuban foredeep (IKB) and the DF. Basement in these areas could be mafic in composition, based on published laboratory data from different regions (Volarovich, 1978; Dortman, 1992; Mooney and Christensen, 1994; Christensen and Mooney, 1996).

The structural complexity of the inferred Palaeozoic succession just to the south of the RD and the faulting within DF are interpreted to be caused by compressional tectonics. There is no clear indication of faulting penetrating the entire crust along Ilskaya–Leningradskaya DSS profile. However, the presence of northward dipping fault zones or structural contacts, forming the boundaries between the high velocity crystalline crust of IKB and the Palaeozoic crust of SP and between the crust of the SP and RD is suggested. The DF was a rift in the middle to late Palaeozoic (Stovba and Stephenson, 1999, Kutas and Pashkevich, 2000). The high reflectivity in the lower crust and below the Moho in the DF along Bataisk–

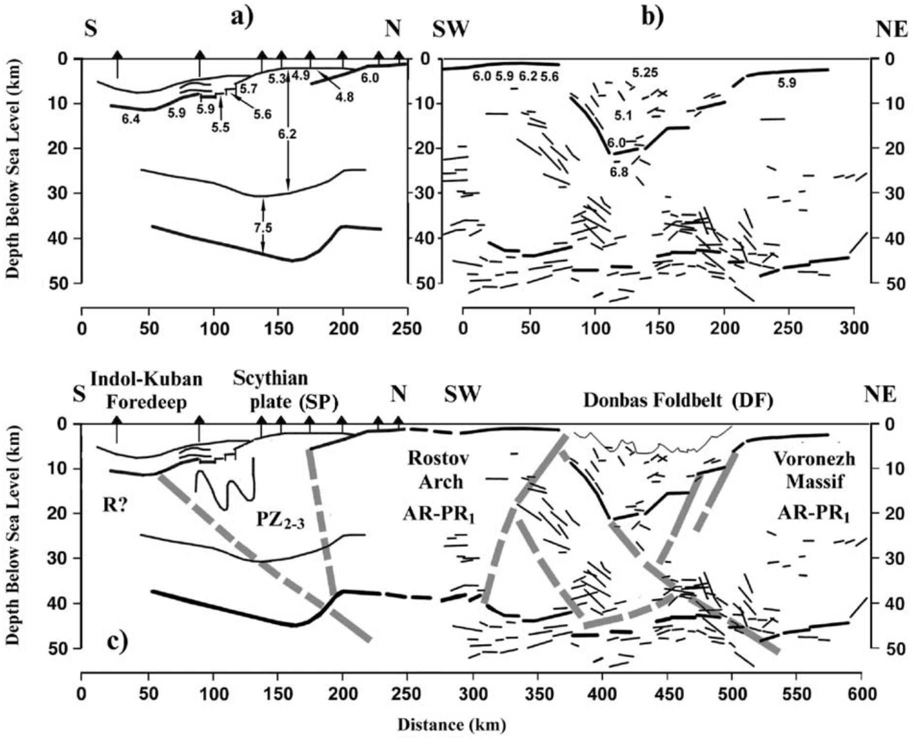


Fig. 2. Crustal cross-sections from the (a) Ilskaya–Leningradskaya (located as 1 in Fig. 1) and (b) Bataisk–Milyutinskaya (located as 2 in Fig. 1), DSS profiles and (c) an interpretative composite cross-section of the crust of the IKB, SP, RD, DF and VM. The numbers in the sections indicate values of P-wave velocity (km/s). Thick solid lines show the basement and Moho boundaries. Thin lines represent reflectors in the crust. Boreholes near the profile are shown for the SP and the southern slope of the RD. The folded and faulted complexes of the crust of SP and of DF are schematically displayed. Thick grey lines represent inferred deep faults, dashed where the interpretation is considered to be most speculative.

Mylyutinskaya DSS line indicates sub-horizontal shear bands, typical beneath a rift (cf. DOBREfraction '99 Working Group, 2003; Maystrenko et al., 2003). A steep fault at the northern margin of the rift with a southward inclination is inferred to descend to about 20 km, cross the entire crust and merge with the shear band below the Moho under the southern flank of the rift. In spite of this, the minor reflectors within the southern portion of the crust beneath the rift indicate a northward inclination (Fig. 2b,c). One explanation is that the southward inclined fault was formed during the extension (rifting) stage with the northward dipping shear band forming during subsequent inversion (shortening).

Fig. 3 displays a model of the crust along the Krasnodar–Emba DSS profile acquired by GEON in 1990–1991 (profile 3 in Fig. 1). It is based on three-component analogue data, spaced every 7–12 km, recording shots of up to 3 tonnes. The profile crosses, from west to east, the IKB, the StH, the KS and the southern portion of PCB in a direction that is not perpendicular to the strike of these tectonic units. Seismic boundaries in the section have dip angles less than what has been observed in the N–S direction. Several fault zones with northward inclination are inferred; these originate within middle and lower crustal low velocity zones and pass through an area of 8.0 km/s velocity in the upper mantle just beneath the Moho. These structures can also be interpreted as collision signatures. A near-vertical offset of the top of the crystalline crust from 3 to 7 km depth between the IKB and the StH indicates a fault in the upper crust. The seismic data also suggest that there are the steep faults on the southern side of the KS that dip

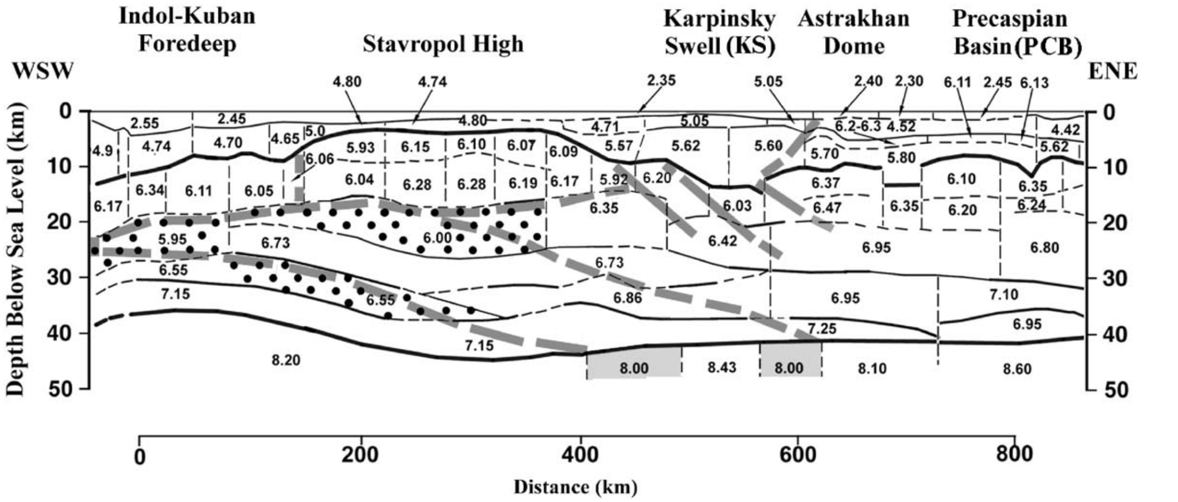


Fig. 3. Crustal cross-section along the Krasnodar–Emba DSS profile (located as 3 in Fig. 1). Thick solid lines show the basement and the Moho and thinner black lines indicate wide-angle reflecting horizons in the crust recognised from P- and S-seismic waves. Thinnest lines are seismic boundaries determined from only one type of seismic wave. Dashed lines are interpolated boundaries. The numbers indicate values of P-wave velocity (km/s). Dots indicate the low-velocity zone. Light-grey colour indicates upper mantle with a P-wave velocity of 8.0 km/s. Thick grey lines represent inferred deep faults, dashed where the interpretation is considered to be most speculative.

northwards to a depth of at least 20 km. At the northern side of the KS (right side of Fig. 3), a south-dipping reverse fault is suggested from coincident CDP data (Brodsky et al., 1993; Brodsky and Voronin, 1994; Volozh et al., 1999). There are low velocity zones at a depth of about 20 km beneath the StH and at the depth of about 30 km under the transition between the IKB and the StH.

Any structural or kinematic interpretation of these velocity data is speculative given a general lack of geological constraint observable at the surface. However, Khain and Sokolov (1991) suggested that the rocks comprising the KS have been displaced in a northward direction by more than 100 km. In this context, it is possible that the StH formed a rigid block that pushed the KS north like a bulldozer, detached from and moving over the low velocity zone in the middle of the crust as an allochthonous unit. As such, the unit just to the south of the StH would have been moving northward as well and the low velocity zone at a depth of about 30 km there could be evidence of a zone of ductile shearing in the lower crust.

Combined DSS-CDP seismic observations along the Morozovsk–Manich profile (4 in Fig. 1) are being carried out at the present time by GEON. Some 350 km of DSS and converted wave recordings as well as 320 km of deep CDP data were acquired to the beginning of 2001. In 2002, a further 650 km of DSS and 450 km of CDP profiling were completed. During the 2001 field experiment, more than 170 "Delta-Geon" digital four-component seismic recorders, spaced every 2–3 km, were used for DSS acquisition and the resulting data are now being processed and analysed. Two Sercel (SN-388) seismic stations (one from GEON Centre and another one from "Spetsgeofizika") were used for CDP acquisition. Spread length was 10 km, using groups of 12 recorders spaced every 50 m. Five 10-tonne vibrators provided the source signal, every 100 m, giving a nominal fold of 50. Record length in the study area was 25 s. Processing was done at the GEON Centre. Fig. 4a shows the CDP section for a fragment of this profile where it crosses the Volgodonsk–Elistian portion of the KS (between 41°30′ E and 45°E on Fig. 1). The Moho discontinuity is at 12–13 s (equivalent to about 40–42 km depth) beneath the northern portion of KS deepening to 14–16 s (~ 46–48 km) below its southern margin of KS (Fig. 4a,b). High-energy reflectors with southward inclination are observed in the middle crust between 4 and 10 s TWT. These may be related to the

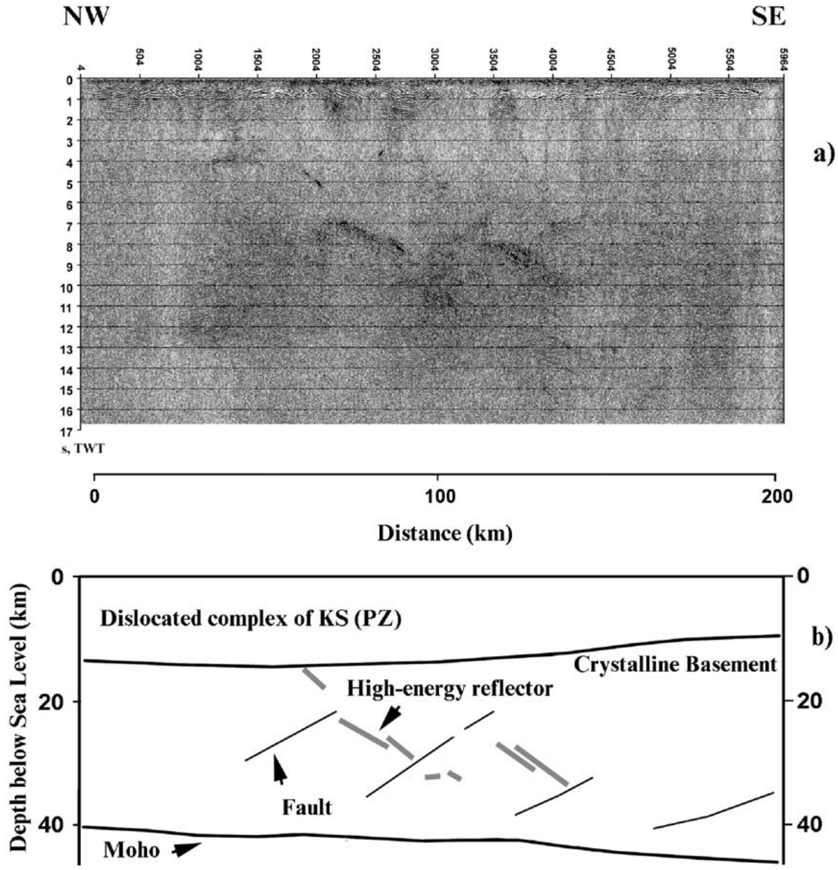


Fig. 4. (a) CDP stack and (b) its interpretation across the KS along Morozovsk–Manich profile (thick portion of line 4 in Fig. 1). The position of the boundary between the Palaeozoic (PZ) complex and crystalline basement is derived from coincident DSS data.

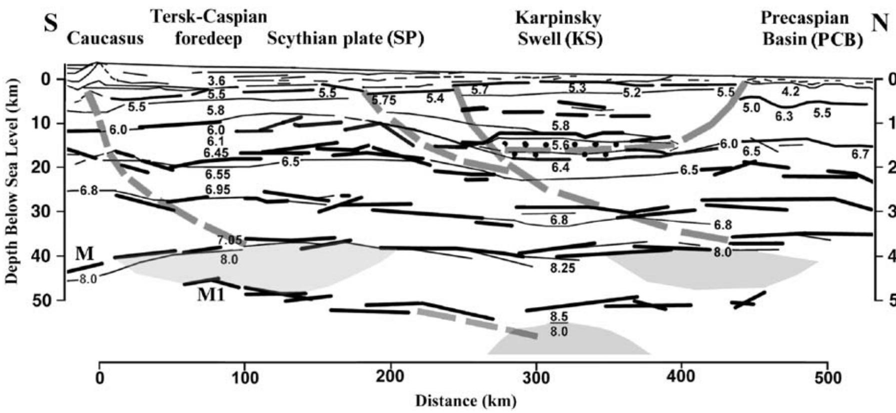


Fig. 5. Crustal cross-section along Volgograd–Nakhichevan DSS line (located as 5 in Fig. 1). Thick lines are wide-angle reflectors. Lines of intermediate thickness are boundaries based on refracted seismic phases. Thin lines are P-wave isovelocity contours. Velocities are given in km/s. M and M1 horizons are discussed in the text. Thick grey lines represent inferred deep faults, dashed where the interpretation is considered to be most speculative.

internal structure of the crystalline crust below the KS although individually they may correspond to fault surfaces. Other reflectors with opposite polarity are seen in the middle and lower crust. These are interpreted as northward dipping faults (Fig. 4b). The position and orientation of these reflectors coincide with the northward inclined faults inferred from the DSS velocity model seen in Fig. 3 and, as such, they are interpreted to be related to convergent processes on the KS, as it was displaced from south to north.

Fig. 5 shows a velocity model along the Volgograd–Nakhichevan DSS profile (profile 5 in Fig. 1) from Krasnopevtseva (1984). From south to north, this profile crosses the Tersk–Caspian foredeep (TCB), the SP, the KS, and the western part of the PCB. There is a low velocity zone below the 5.2–5.8 km/s complex of the KS just above what is interpreted as the top of crystalline crust (e.g. 6.4 km/s). This zone could represent the results of tectonic layering or faulting developed, for example, during the northward displacement of the KS. Further, a northward dipping fault is inferred that cuts the crust beneath the KS from the top of basement to the Moho reaching the upper mantle where the seismic velocity is about 8.0 km/s. There is also an upper mantle horizon, based on a wide-angle reflected seismic phase, dipping from ~45 km under the

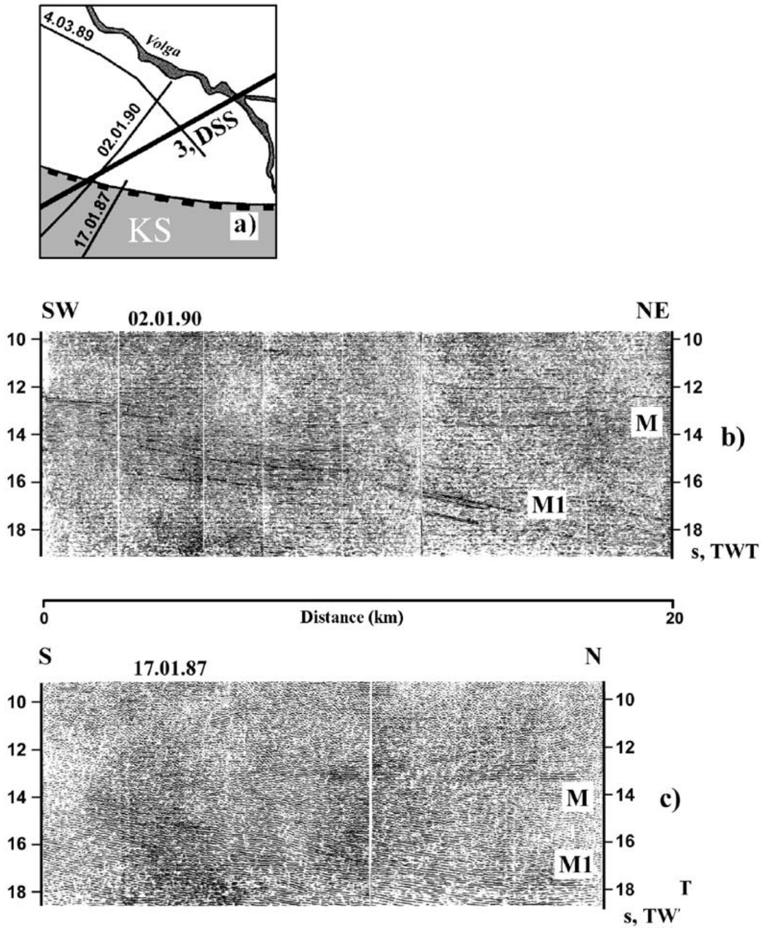


Fig. 6. (a) Location map (indicated as frame 6 in Fig. 1), (b) CDP stack 02.01.90 and (c) CDP stack 17.01.87 from the Astrakhan dome area. M = Moho boundary, M1 = subducting slab in the upper mantle.

TCB to >52 km beneath the southern margin of the KS. Between the Moho and this horizon, the P wave velocity varies from 8.0 to 8.25 km/s. This geometry possibly indicates a doubling of the Moho, a crustal underplating, that might be linked to an environment of subduction.

Near vertical seismic observations were acquired in the transition between KS and PCB (Figs. 1 and 6a) by the Astrakhan Geophysical Expedition in 1987–1990 (Brodsky et al., 1993, Brodsky and Voronin, 1994). A "Progress" seismograph was used; spread length was 4.8 km (and sometimes 9.6 km), with groups of recorders and shots every 50 m, giving a nominal fold of 48. Record length in the study area was 20 s. Fig. 6b and c displays fragments of lines 17.1.87 and 2.1.90, respectively. Splitting of the Moho is clearly seen. Brodsky et al. (2001) interpreted two subducting slabs (the first one beneath the KS along line 17.1.87 and the second one beneath the Astrakhan dome along line 2.1.90).

Brodsky et al. (2000) constructed a model cross-section of the Astrakhan dome along line 4.03.89 (Fig. 6a) based on a combination of CDP and conventional refraction data and this is shown in Fig. 7a,b. In the central part of cross-section, the velocity of 6.9 km/s is interpreted to indicate basic magmatic rocks lying beneath a refraction boundary at depth of about 12 km. The velocity along this boundary decreases to 6.3 km/s both to the north and south suggesting that the inferred basic rocks are limited to a zone about 40 km wide. Inclined reflectors in the depth range 12–24 km are thought to define the shape of proposed magmatic body. Sediments underlying the Devonian succession inferred to be of Ordovician–Silurian age (Fig. 7b) directly overlie the magmatic body, although the age of these sediments is not known for certain. If they are of Ordovician–Silurian age, then older, pre-Ordovician volcanic units may have existed in the area of the more recent Astrakhan dome.

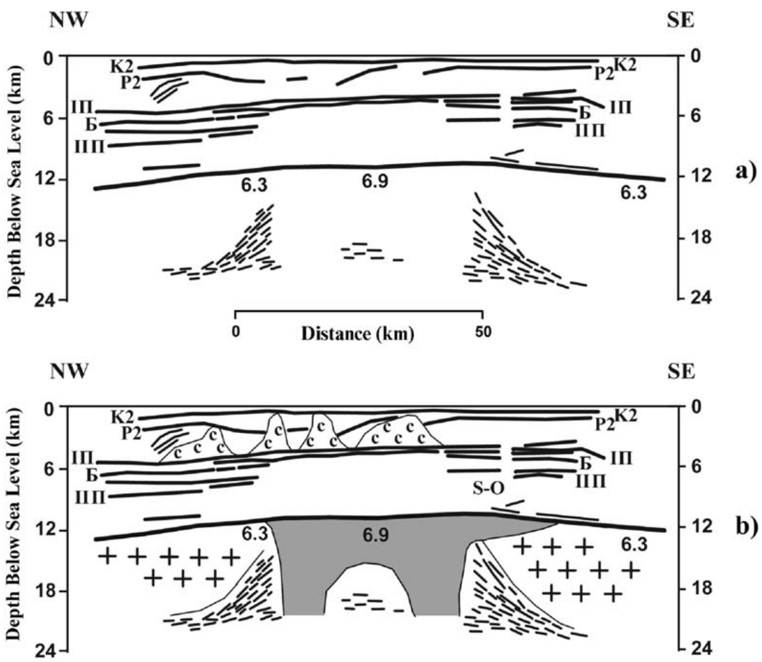


Fig. 7. (a) Seismic cross-section and (b) a geological interpretation of the upper crust of the Astrakhan dome along CDP profile 4.03.89 (located in Fig. 6a). Thick black line indicates the basement surface, based on refraction data. Lines of intermediate thickness are seismic boundaries within sedimentary cover and the short thin lines are reflectors in the basement. Numbers indicate values of P-wave velocity (km/s). Regionally identified seismic boundaries are as follows: K2 and P2—in the Permian to Cainozoic successions, IIП—in middle Carboniferous strata; Б—in Lower Carboniferous strata and IIIП—in middle Devonian rocks. Salt-related structures are shown by the c-symbol; S–O indicates Silurian to Ordovician sequences; crosses indicate crystalline basement and grey shading indicates what is interpreted as a mafic magmatic complex.

### 2.2. Gravity field

Yegorova et al. (2004) have presented a three-dimensional analysis of Bouguer gravity anomalies in the northern part of the study area. Here, the Bouguer anomalies, defined on a regular 10 by 10 km grid, are used to determine the second residual component of the gravity field (Fig. 8), which, being rich in shorter wavelengths, gives an enhanced image of smaller scale heterogeneity within the study area.

The VM, the Ukrainian Shield (UkS) and the RD, as well as the SP, KS and PCB are all evident on second gravity residual map. Negative anomalies less than-50 mGal lie along the southern and southeastern flanks of the VM, next to the DF and PCB respectively. The high gravity gradient suggests that the boundary between the VM and these basins is formed by a fault. Positive residual anomalies greater than +40 mGal are seen within the PCB. Seismic interpretations presented above (Fig. 7b) suggests that local positive anomalies (∼ +25 mGal) over the Astrakhan dome could be related to basic magmatic rocks lying beneath the sedimentary cover.

The residual gravity data suggest a sharp border between the KS and the area just to the south of it; however, the northern boundary of the KS, with the PCB, is dissimilar and not distinct. Residual highs (∼ +25 mGal) characterise the eroded DF, while the eastern portion of KS is characterised by a smooth positive field (in the range 0– +20 mGal). In contrast to this, several low amplitude local positive anomalies lie along the southern margin of the western part of the KS (between 41°30′ E and 45°E) with small negative anomalies along its northern margin. These features of the residual gravity field taken together imply that the deep structure of western portion of the KS differs from that of the DF and eastern KS.

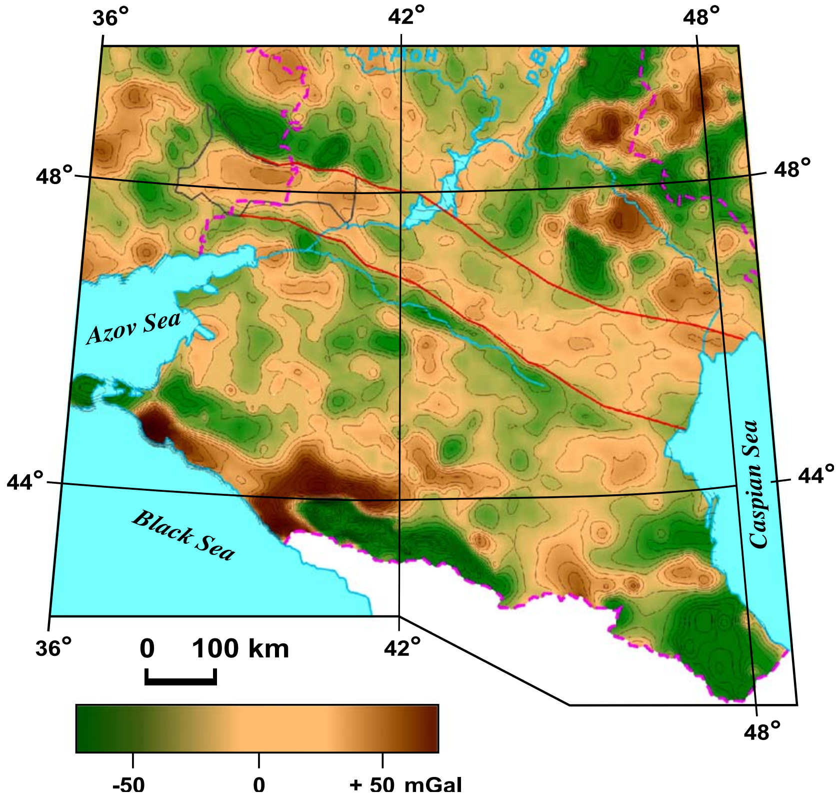


Fig. 8. Second residual component of the Bouguer gravity anomalies in mGal. Black line outlines the eroded Donbas Foldbelt and the red line outlines the Karpinsky Swell. Dashed red line is the boundary between Russia and Ukraine (left) and between Kazakhstan and Russia (right).

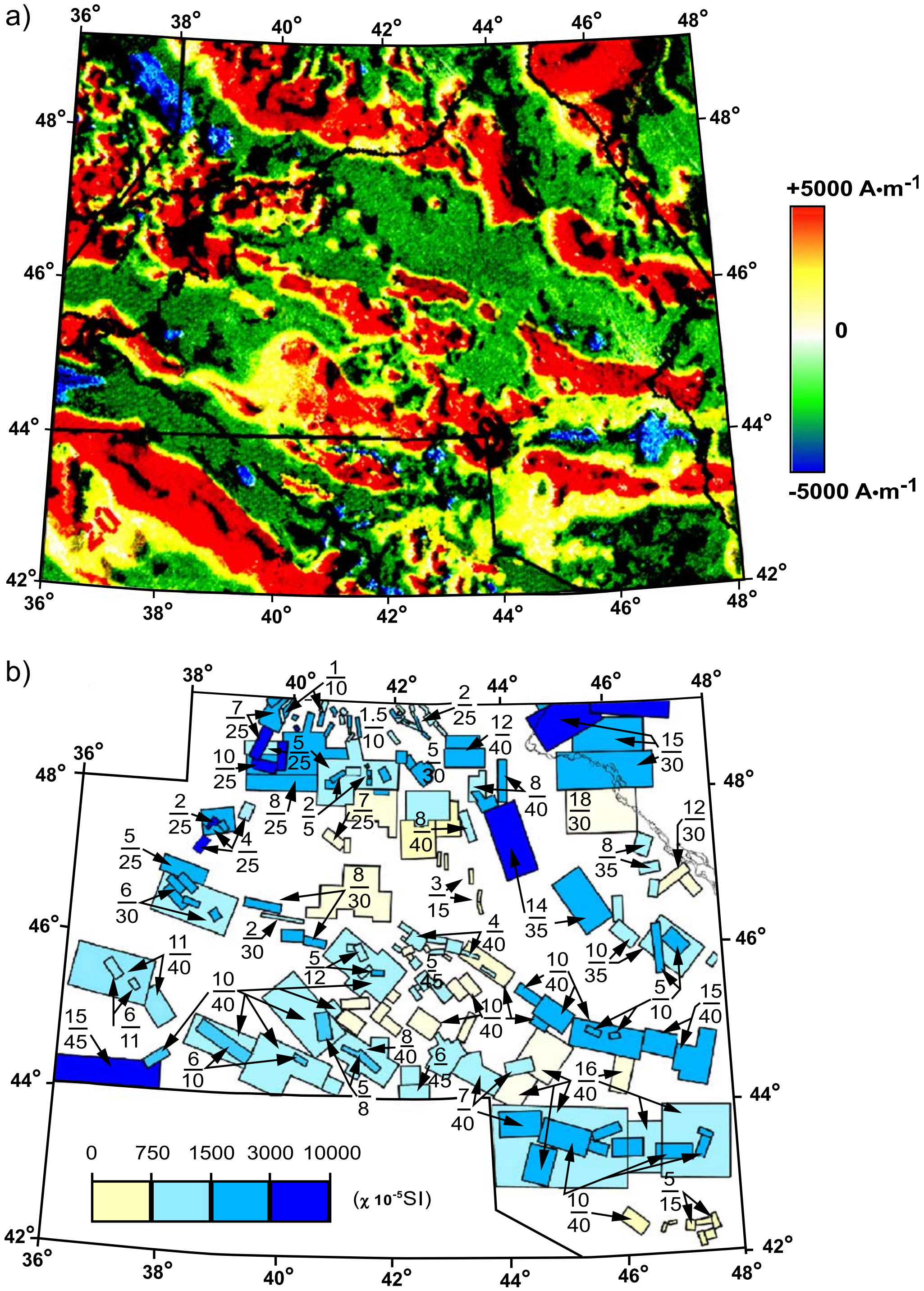


Fig. 9. (a) Magnetic map of the study area and environs based on the 1:5,000,000 map published by Litvinova (2000). (b) Simplified map of the distribution of magnetic bodies determined from 3-D modelling of the magnetic field seen in (a). Numbers indicate depths to the top (numerator) and bottom (denominator) of the model magnetic sources and grey shades indicate the adopted magnetic susceptibility.

The StH and Mineralnie Vody Dome (MV) mostly display positive gravity anomalies up to about +40 mGal (Fig. 1). High gravity gradients and negative residual anomalies (< −20 mGal) suggest the presence of faults along the northern and southern boundaries of MV-StH area.

### 2.3. Magnetic field

Maps of the magnetic field of the study area have been published at scales of 1:2,500,000, 1:5,000,000 and 1:10,000,000 and prepared at scales of 1: 200,000 and 1:1,000,000 by the State Geological Survey of the former Soviet Union. Fig. 9a is based on the 1:5,000,000 map published by Litvinova (2000). A 3-D model of magnetic sources that satisfactorily replicates these observations is seen in Fig. 9b. The actual modelling was carried out a scale of 1:1,000,000 using an algorithm developed by Bhattacharyya (1966). The model consists of a number of independent bodies with uniform magnetisation and simplified shape. The locations and trends of high magnetic gradients were used to determine the plan-view shape magnetic bodies. Their depths were estimated by direct examination of the shapes of associated magnetic anomalies and application Pyatnitsky's "tangent technique" of (Exploration Geophysics, 1964). Only induced magnetisation has been considered and magnetic susceptibilities have been chosen to give the best fit to the observed magnetic field using a trial and error approach.

The 3-D model—showing the locations of magnetic source bodies, depths to their tops and bottoms and susceptibilities—is presented in Fig. 9. Source bodies with the greatest magnetic susceptibilities—up to $10,000 \times 10^{-5}$ SI—are found in the VM, UkS and RD areas, where Precambrian basement, which is assumed to host them, is close to or at the surface. In general, depth to "magnetic basement" is in good agreement with basement depth derived from the seismic data. In particular, source elements within crystalline crust are found along the southern and southern boundaries of the VM near the margins of the DF and PCB. These model elements have, on average, widths of 40–50 km, lengths >100 km, thicknesses of about 20 km and magnetic susceptibilities in the range $1500–5000 \times 10^{-5}$ SI.

A linear zone of positive magnetic anomalies in the magnetic field north of the northern boundary of KS (Litvinova, 2000) is modelled with a series of source elements with their tops at depths from 10 km to 15 km (Fig. 9). Several small bodies are shallower (in the depth range 6.0–3.5 km) and may reflect the influence of magmatic rocks within the Palaeozoic sedimentary complex. Rocks with low magnetic susceptibility ($0–750 \times 10^{-5}$ SI) exist within the eastern DF and western KS, while there are no magnetic sources within the eastern portion of KS. Magnetic bodies, at depths 10–25 km within the crystalline basement, having magnetic susceptibilities of about $1500–3000 \times 10^{-5}$ SI and coinciding with a high gravity field gradient, support an presence of a fault along the southern boundary of the eastern part of the KS. It is possible that it has been intruded by basic magmatic rocks. The inferred magnetic source elements have a uniform magnetic susceptibility to a depth of 40 km, suggesting that such a fault crosses the entire crust from the basement surface down to the Moho.

The 3-D magnetic model of the area of the MV and the StH comprises an ensemble of small source bodies. Their tops occur at depths 2–6 km in the crystalline basement and they have magnetic susceptibility of about $1000 \pm 500 \times 10^{-5}$ SI. There are no geological data about the composition of the basement of the StH. Where the MV is exposed, Letavin (1980) and Milanovsky (1987) described Precambrian to early Cambrian aged metamorphic and magmatic rocks. The potential field data suggest that similar rocks constitute the StH.

## 3. Supracrustal and crustal structure maps

The geophysical data and their interpretations described above have been used to construct maps of the depth to crystalline basement (Fig. 10), the depth to Moho (Fig. 11) and of crustal thickness (Fig. 12). In general, little is known of the crystalline crust of the study area, with the exception of the VM, UkS and RD. Exposures and boreholes in these areas reveal Archean–early Proterozoic complexes comprising igneous and amphibolite to granulite facies metamorphic rocks. Elsewhere, in particular within the DF, KS and PCB, the thick (up to 20 km) Palaeozoic to

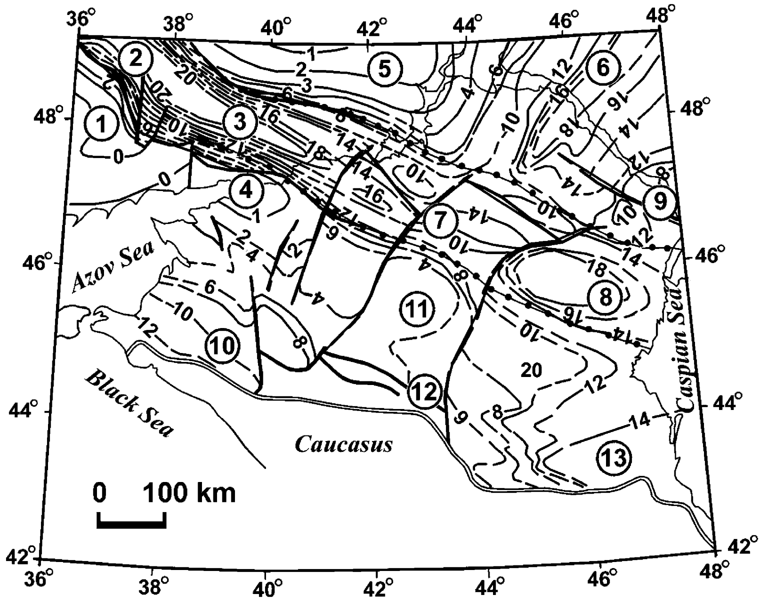


Fig. 10. Depth to basement as discussed in the text. Depth contours (in km) are continuous where well established and dashed where they are interpolated. Thick line shows the fault. Thin dotted line indicates the recent shape of the Karpinsky Swell. Doubled line displays the boundary of Caucasus. Numeric legend: 1 = Ukrainian Shield, 2 = Dniepr–Donets Rift, 3 = Donbas Foldbelt, 4 = Rostov High, 5 = Voronezh Massif, 6 = Precaspian Basin, 7 = Volgodonsk–Elistian portion of the Karpinsky Swell (KS), 8 = Tsubuk–Promislovian portion of KS, 9 = Astrakhan dome, 10 = Indol–Kuban foredeep, 11 = Stavropol High, 12 = Mineralnie Vody Dome, 13 = Tersk–Caspian foredeep.

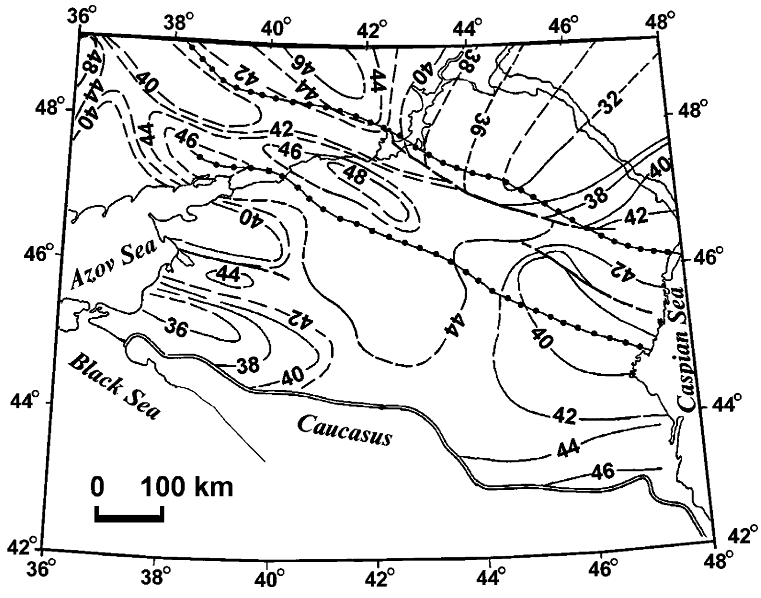


Fig. 11. Depth to Moho (km); legend as in Fig. 10.

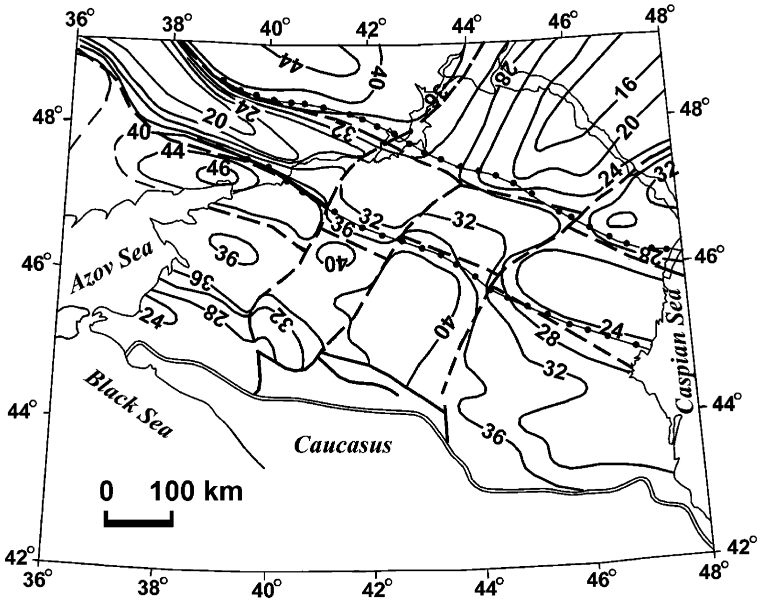


Fig. 12. Thickness of crystalline crust (km). Dashed lines shows boundaries between what are inferred to be different crystalline crustal blocks; otherwise the legend is the same as for Fig. 10.

Cainozoic sedimentary successions overlying crystalline basement has ensured that nothing certain is known of its composition and age.

The top of crystalline basement (Fig. 10) is defined as corresponding to the refraction seismic boundary with velocity of greater than 6.0–6.2 km/s as determined from DSS and conventional seismics. On the basis of this definition, the depth to basement within the VM varies in the range 0–4 km. In the DDB and DF, it reaches 20 km. In the PCB, the basement surface is also as deep as 20 km, but deepening is much more gradual, occurring over a distance of about 100–130 km compared to the DDB and DF where the basement surface deepens very abruptly. The western (Volgodonsk–Elistian; 41°30′ E to 45°E) segment of the KS has a basement depth in the range 10–16 km, whereas in its eastern (Tsubuk–Promyslovian; 45°E to 48°E) segment it is deeper (18–20 km).

The SP area comprises three segments: the IKB in the west, the StH in the centre and the TCB in the east. Within the StH, boreholes penetrated a late Devonian–early Carboniferous terrigenous succession at a depth of about 1.2–2.0 km, and the seismic data indicate crystalline basement at 4.0–5.0 km. Within the IKB and TCBs, crystalline rocks are poorly defined by the seismic data at a depth of about 10–12 km.

Fig. 11 shows depth to Moho; seismic data in the SP region in particular are sparse and the Moho depths as shown can be considered to be schematic only. Moho depth varies in the range 32–46 km. The VM, the UkS and the RD have a crust of about 44–46 km thick. Moho on the flanks of the PCB is at a depth of 40–42 km shallowing to 32–36 km in its central part. Moho depth beneath the DF is not greater than about 40 km according to newly published results (DOBREflection-2000 and DOBREfraction '99 working groups, 2002; DOBREfraction '99 Working Group, 2003), which have incorporated here. This compares to older DSS studies (in which Moho depth was based primarily on reflected seismic phases) where Moho was placed at about 48–50 km (Sollogub et al., 1978; Sollogub and Chekunov, 1980; Kutas and Pashkevich, 2000). In the axial part of the western KS, the Peschanokopskaya–Surovikino DSS data indicate a Moho depression to 48 km (Konovaltsev et al., 1980). This is partially supported by the new Morozovsk–Manich CDP data (Fig. 4a,b), which show that Moho lies at about 40–42 km beneath the northern margin of the KS and deepens to 48–50 km to the south. This contrasts with 40–42 km in the eastern KS based on the Nakhichevan–Volgograd DSS profile (Fig. 5).

The Nakhichevan–Volgograd DSS profile also indicates that, within the eastern portion of the SP, the Moho shallows from 46 km under the pre-Caucasus foreland to 40 km within the SP-KS transition. Under the IKB, the Ilskaya–Leningradskaya DSS profile (Fig. 2a,c) indicates a Moho depth of about 38 km. Under the StH the Moho lies at a depth of about 44–46 km based on the Krasnodar–Emba DSS profile (Fig. 3).

Fig. 12 displays the thickness of the crystalline crust in the study area and, as such, is simply the Moho depth (Fig. 11) less the depth to crystalline basement (Fig. 10), calculated on a regular 25 by 25 km grid. The VM, the UkS and the RD have very thick crystalline crust, about 40–44 km, consistent with their affinity with the EEC. In contrast, crystalline crust is only 14–16 km thick in the central PCB, explained by Riphean aged rifting by Kiryukhin et al. (1993). Brunet et al. (1999) suggested active rifting phases in Riphean time as well as during the Vendian–Ordovician (although this is poorly constrained) followed by late Devonian subsidence in an extensional setting possibly due to back-arc rifting. Nevolin and Fedorov (1995) described the Precambrian stage of PCB evolution in terms of a triple junction with the Uralian paleo-ocean (east), the Embinian aulacogen (south) and the Pachelma aulacogen (north). Kostyuchenko et al. (1999) invoked Riphean and middle Palaeozoic stages of rifting to explain the crustal structure of the PCB. The crystalline crust of the DF is about 20–22 km thick, which is very similar to that of the DDB. Crustal thinning in these areas is entirely due to late Palaeozoic rifting according to Stovba et al. (1996) and Stovba and Stephenson (1999). In the western KS, the thickness of crystalline crust varies from 28 km to 34 km, significantly greater than in the DF; in the eastern KS it is less than 24 km. Taken together, these data do not support the hypothesis of Volozh et al. (1999) that the DDB, DF and KS constitute a single intracratonic rift with uniform crustal parameters.

The StH has a crystalline crustal thickness of about 40 km. In contrast, the crust of the IKB is less than 28 km and beneath the TCB it decreases from 36 to 32 km thickness from south to north. Such a significant difference in crustal thickness suggests that the crust of the StH has a different crustal affinity than the rest of the SP.

## 4. Summary and discussion: geodynamic framework and evolution of the southern margin of the EEC

### 4.1. Key interpretations

Given the generally poor state of knowledge of the geology of the basement in the study, the geophysical data and their interpretations presented above constitute the main constraints understanding the crustal structure and tectonic evolution of the southern margin of the East European Craton and adjacent regions. The recognised lateral and deep variations in the crustal and upper mantle structure of the study area provide a basis for a critical reappraisal, in a plate tectonic framework, for existing ideas on its tectonic history—although necessarily with a degree of speculation. In so doing, several key interpretations are utilised and they are as follows:

1. The UkS, the RD and the southern part of the VM are the peripheral elements of the EEC, as is conventionally understood. The Archean–early Proterozoic age of the crystalline crust, covered by Palaeozoic to Cenozoic sedimentary sequences, and its thickness of up to 46 km constitute the major characteristics of these tectonic units.
2. The DF is a thrust-faulted and folded, inverted, southernmost segment of the DDB, which formed as an intracratonic rift in the middle and late Devonian (Stovba and Stephenson, 1999; 2000). There is a widespread early Permian unconformity, which is much more pronounced on the southern margin of the DF than on its northern one. Based on this and other considerations, Popov (1963) and many others suggested that the major tectonic phase leading to the development of compressional deformation in the DF occurred during the latest early Permian. Stovba and Stephenson (1999), however, based primarily on newly acquired regional seismic reflection profiling, concluded that this unconformity related to the uplift of the southern flank of the basin and of the neighbouring

UkS under a trans(tensional) tectonic regime rather than to crustal shortening. They argued that compressional deformation took place in latest Triassic times and, mainly, in latest Cretaceous–earliest Tertiary times. Interpretation of the Bataisk–Milyutinskaya DSS profile (Fig. 2b,c) suggests that south dipping faults were formed during rifting while a north dipping shear zone formed during shortening.

3. The crust of the KS, underlying the Mesozoic–Cainozoic sedimentary cover, consists of three seismic layers (cf. Krasnodar–Emba DSS; Fig. 3). P-wave velocity varies in the range 4.7–5.05 km/s in the upper layer, which is 3.0 km thick. Borehole data provide some evidence that middle Palaeozoic terrigenous rocks constitute this layer. The middle layer is 12–13 km thick and has a P-wave velocity of about 5.6 km/s. There is no borehole penetration of this layer. The lower layer is crystalline crust, in which velocity increases with depth from 6.0–6.15 km/s at the top to 7.15 km/s at the base. As such, the architecture of the KS crust is very similar to that of Honshu Island (Murauchi and Yasui, 1968), Lord Howe Rise (Shor et al., 1971) and the Kuril'skiye Islands (Tuezov, 1975). It also compares closely to the average crustal structure of "continental arc" as determined by Nikolas and Mooney (1995) and Mooney et al. (1998). The northern and southern boundaries of the eastern KS is characterised by a linear zone of positive magnetic anomalies while a negative magnetic field is characteristic for the rest of the KS. This pattern is similar to the magnetic field above the Kuril'skiye volcanic islands (Litvinova, 2000). Pre-Ordovician, possibly of Riphean age, magmatic bodies are evident in the 04.03.89 CDP seismic profile (Fig. 7a,b). Evidence of a subducting slab, dipping northwards beneath the proposed volcanic arc, is seen on the 17.01.87 CDP profile (Fig. 6b,c). It can accordingly be speculated that the eastern part of the KS consists of volcanic arc.
4. Crustal shortening and thrusting are suggested by compressional signatures seen in the western part of the KS along Morozovsk–Manich seismic profile (Fig. 4a,b).
5. The presence of pyroclastic rocks of variable composition in the DF, DDB and PCB indicates volcanic activity during the early and middle Carboniferous (Kalashnikov, 1974; Vishnevskaya and Sedaeva, 2000). Furthermore, spicul-and radiolaria-bearing sediments documented by Kireeva and Maksimova (1959) and Vishnevskaya and Sedaeva (2000) indicate that a marginal or back-arc marine basin existed just to the north of the KS and in the area of the DF during middle Palaeozoic times.
6. The upper crustal unit beneath the StH may be detached from the underlying crust across a significant low velocity zone at a depth of 20–30 km (Fig. 3). As such, it may have moved northwards independently of the lower crust indenting the back-arc basin ensemble of what is now the western KS by as much as 100 km, as argued by Khain and Sokolov (1991).
7. Northward dipping faults inferred in the consolidated crust of the SP on the Ilskaya–Leningradskaya (Fig. 2c), Krasnodar–Emba (Fig. 3) and Volgograd–Nakhichevan DSS profiles (Fig. 5) may be related to compressional structures seen in the KS and, if so, argue that the Caucasus area was involved. However, there is no age constraint on the formation of these structures. In any case, they cannot be older than early Permian and given the age of compression on the KS and DF, they formed at least in part during Cimmerian (Triassic–Jurassic) and/or Alpine (Cretaceous/Tertiary) tectonic episodes.

### *4.2. Schematic tectonic model*

A schematic model of the evolution of the southern margin of the EEC and adjacent regions, based on the observations and interpretations summarised above, is presented in Fig. 13.

The Astrakhan Dome, which is the westernmost unit of North Caspian–Aktyubinsk zone of basement uplifts in the PCB (Brunet et al., 1999; Volozh et al., 2001), is speculated to represent a Riphean-aged (?) volcanic arc (Fig. 13a). In any case, it is older then Ordovician. Accordingly, the Caucasus–Peredneasian basin of the Proto-Tethys Ocean (Khain and Rudakov, 1996; Rudakov, 2000) could have existed just south of this arc. Back-arc extension occurred in the area of Precaspian Basin, resulting in the earliest phase of its subsidence (cf. Brunet et al., 1999).

Fig. 13b corresponds to the time of formation of the DDB/DF rift. The RD, on the southern flank of the DF, moved southwards as a result of extension in a paleo-Scythian back-arc basin. The middle Palaeozoic rock ensemble of what is now the western (Volgodonsk–Elistian) segment of the KS would also have been further to the south. The eastern (Tcubuk–Promislovian) segment of the KS, however, as determined by the linear zones of positive magnetic anomalies, continued to develop as a volcanic arc. The important late Palaeozoic extensional phase in the PCB took place.

Fig. 13c summarises the ensuing phases of (latest early Permian?) Mesozoic and Cenozoic compressional tectonics and shortening of the paleo-Scythian crust leading to the assembly of the present crust of the area and the Caucasus. Folding and faulting of sedimentary successions in the area, the displacement of the RD to the north, and formation of the DF all would have occurred during this time. The StH formed an indentor that moving to the north, displacing the middle Palaeozoic, mostly sedimentary, complex by more than 100 km.

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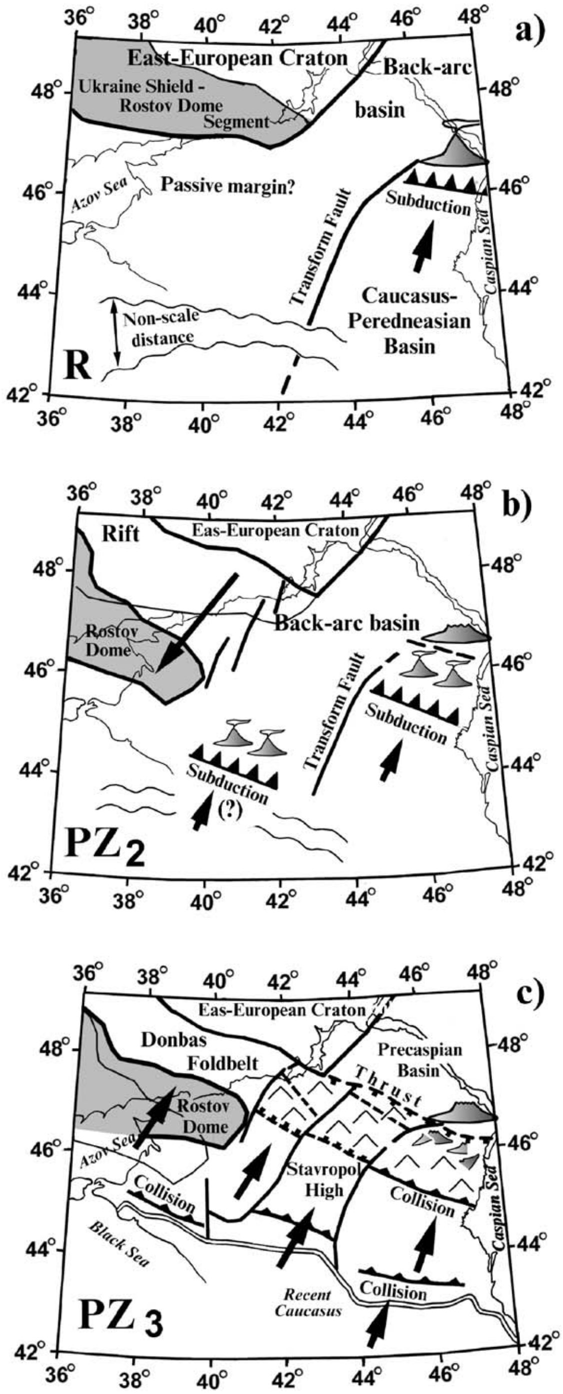


Fig. 13. Schematic model of the tectonic evolution of the study area: (a) Precambrian, possibly Riphean; (b) middle to late Palaeozoic time; and (c) (latest early Permian?) Mesozoic–Cainozoic time. Grey colour indicates the Ukrainian Shield–Rostov Dome segments of the East European Craton; the thin black line in (b) and (c) displays the earlier positions of this unit. Active and non-active volcanic arcs are indicated by volcano and eroded volcano symbols respectively. Volcano fragments indicate that an arc has been involved in accretionary tectonics. Angles show assemblages of rocks deformed during collision. Black lines display transform and strike-sleep faults; dashed lines indicate normal faults, reverse faults and thrusts. Arrows point the direction of displacement of crustal units.

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